

A dry Venusian interior constrained by atmospheric chemistry

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Venus's climatic history provides powerful constraints on the location of the inner edge of the liquid-water habitable zone. However, two very different histories of water on Venus have been proposed: one where Venus had a temperate climate for billions of years with surface liquid water and the other where a hot early Venus was never able to condense surface liquid water. Here we offer a constraint on Venus's climate history by inferring the water content of its interior. By calculating the present rate of atmospheric destruction of H₂O, CO₂ and OCS, which must be restored by volcanism to maintain atmospheric stability, we show that Venus's interior is dry. Venusian volcanic gases have at most a 6% water mole fraction, which is substantially drier than terrestrial magmas degassed at similar conditions. The dry interior is consistent with Venus ending its magma ocean epoch desiccated and thereafter having had a long-lived dry surface. Volcanic resupply to Venus's atmosphere, therefore, indicates that the planet has never been liquid-water habitable.

Venus's geodynamic and climatic history is uncertain, particularly regarding whether its surface ever hosted habitable conditions. At its heart, this is a question of whether liquid-water oceans have ever existed on Venus. Unlike Mars, for which spectacular mapping of its surface has revealed its sculpting by water¹, Venus's global resurfacing at least ~0.3 billion years ago has obscured much of its ancient geological history². One observation in support of water having existed in some form on Venus is its large deuterium/hydrogen (D/H) ratio, which is 150 ± 30 times that of Earth³. This observation has been used to infer that Venus once had a surficial water reservoir of similar mass to Earth's oceans⁴: equivalent to up to 500 m of global water depth on Venus^{5,6}. Alternatively, the observed D/H ratio could be the result of a steady-state supply. In this scenario, water is continuously delivered through cometary material or volcanic outgassing, balancing loss through H escape⁷. Critically, the atmospheric D/H ratio alone does not constrain the existence of liquid water on Venus, regardless of what it implies about the quantity or source of water. Hence, the past surface and climatic conditions of Venus are still unknown.

In the absence of definitive observations, climate modelling has been crucial in informing our picture of early Venus, from which two contrasting paradigms emerge: 'temperate and wet Venus' and 'dry

Venus' (Fig. 1). In the temperate scenario, Venus had an Earth-like past. A long-lasting period of temperate climate followed the planet's formation, with shallow oceans (for example, refs. 8,9). General circulation models demonstrate that such a state could be maintained if Venus was initialized cool, with liquid water at its surface, by water clouds that form on its dayside^{10,11}. A cloudy dayside and clear nightside reduce radiation absorption and enhance efficient re-radiation to space, keeping surface temperatures cool enough for water to condense. From this habitable state, Venus transitioned to its present CO₂ greenhouse during the 'great climate transition', probably due to a massive outgassing of CO₂ and SO₂ during the large igneous province volcanism^{11,12}.

In contrast to this is the dry Venus scenario. Here, Venus became desiccated early in its evolution due to the slow solidification of its magma ocean over ~100 Myr (ref. 13). The long-lived steam atmosphere would have enabled dissociation and loss of water through the intense hydrodynamic escape of H (ref. 14), which would have extended throughout the whole planetary interior¹³. With subsequent volcanic outgassing, Venus would have evolved towards its current CO₂- and N₂-dominated atmosphere. This scenario is also supported by a general circulation model, which suggests that an initially hot Venus without surface liquid water underwent nightside water cloud formation¹⁵.

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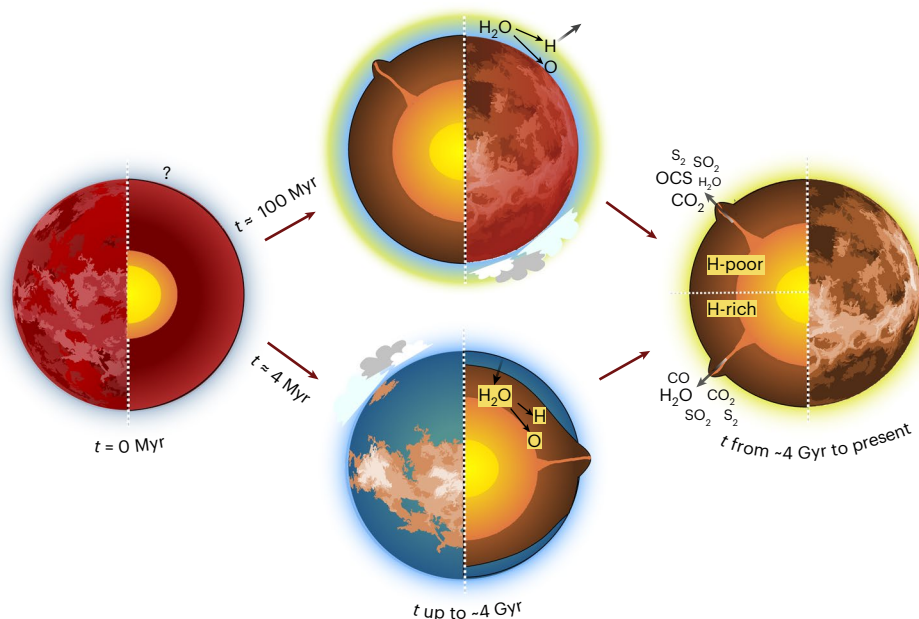


Fig. 1 | The dichotomous climate pathways proposed for Venus. Venus emerges from its formation in a magma ocean phase ($t = 0$ Myr). From this point, Venus could have followed one of two possible pathways, each ultimately converging on the planet's present climate state but leaving a mantle with a different composition: an H-poor interior results from the dry Venus path (upper) and an H-rich interior from the temperate and wet Venus path (lower).

In the H-rich interior scenario, volcanic gases are enriched in H_2O , whereas in the H-poor case, degassing of H_2O is minimal, with a greater release of S and C species. See the text for a detailed description of climatic paths and 'Signs of climate in the interior' for an explanation of how H_2O oceans are imprinted in the planetary interior.

These nightside clouds had a strong net warming effect (by blocking the radiation of dayside heat to space) and prevented surface water condensation even at moderate insolation¹⁵. Such a scenario predicts that an ocean-free early Venusian surface would remain ocean-free thereafter. Any water preserved in the hot atmosphere after solidification of its magma ocean may then have been slowly lost over the planet's history¹⁶.

These dichotomous predictions of Venus's evolution are illustrated in Fig. 1. The essence of the different outcomes of the general circulation models is whether the models are initialized hot (in which case the planet cools inefficiently¹⁵) or cold with an ocean (in which case it can remain temperate for geological timescales¹⁰). A key question then is whether Venus could have cooled sufficiently following the high-energy conditions of planetary assembly (Fig. 1, left) to allow liquid water to form on its surface (the lower path of Fig. 1) or whether a hot-start forced early water loss (the upper path of Fig. 1) and a permanently dry Venusian surface¹⁷.

To distinguish between the dichotomous climate scenarios demonstrated in Fig. 1, we present an approach informed by observations of Venus's current atmospheric chemistry. We calculated what the net composition of volcanic resupply to Venus's atmosphere must be to maintain the atmospheric steady state at the present day. This constrains how wet the planet's interior is, which serves as the source of volcanism. Therefore, we can constrain the planet's internal water inventory, which is distinct between the two dichotomous climate histories.

Signs of climate in the interior

To determine if water once existed on Venus's surface in the form of oceans—indicating traditional liquid-water habitability—we investigated whether present-day Venus has a water-rich interior. We suggest that a habitable past will be associated with Venus's present interior being water-rich and a dry past with Venus's present interior being dry.

Most of Venus's water inventory was probably delivered during the main accretion phase and not during late accretion¹⁸. This makes Venus's magma ocean stage key in setting the initial water inventory

and distribution on the planet. Slow solidification of the magma ocean largely depletes a planet's water inventory. Depletion can occur down to less than a tenth of Earth's ocean mass, even when starting with a water budget ten times larger¹³. Following crystallization of the magma ocean, a substantial proportion of the water left partitions into, and remains within, the planet's mantle^{19–22}.

The above discussion indicates that if Venus's water inventory had been depleted during the magma ocean stage, then its surface would always have been dry thereafter. In this scenario, volcanism would have provided a one-way path for water out of the planet's interior. Water in the atmosphere would then dissociate into H and O, and the H would hydrodynamically escape the planetary system¹⁵.

If Venus was once habitable (that is, it had surface oceans), however, we would expect a water-rich system, necessitating the rapid solidification of its magma ocean to minimize H loss to space following water condensation¹³. The dry Venus scenario is, therefore, linked to an H-poor interior, whereas the temperate and wet Venus scenario is linked to an H-rich interior, drawing a direct connection between the current interior water reservoir and the past behaviour of the water in the atmosphere (Fig. 1).

If Venus had retained a substantial water reservoir following its magma ocean phase, then much of that water should still be present in its interior. The maximum time window between modern-day Venus and a past Venusian climate that could have supported surface oceans is 3 Gyr (ref. 11). Stagnant lid outgassing rates are probably inefficient²³ and suggest a maximum of 0.4 Earth oceans degassed over this time span²⁴, which is less than half the probable water inventory of a temperate Venus¹⁰. A temperate Venus would also have diminished water degassing due to high-pressure eruption under surface oceans²⁵, further increasing the likelihood of a wet past being reflected by a wet mantle. Noble gas tracers of volcanic degassing rates point directly to the planet having had time-integrated inefficient degassing compared to Earth²⁶. Perhaps only 24% of the radiogenic ^{40}Ar in Venus's mantle has been released²⁷. Limited Ar loss points to a similarly inefficient loss of water, thus preserving evidence of a wet past in its interior.

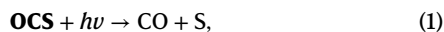
Information on Venus's present interior is afforded through its volcanism. Volcanism on rocky planets is overwhelmingly supplied by partial melting of their hot convecting mantles. Elements such as H are typically incompatible during these melting processes^{28,29} and are, therefore, efficiently transported towards the lower pressure surface where they degas into the atmosphere. Most of the mantle water inventory is in the upper mantle³⁰, where it can depress the melting temperature of mantle rocks³¹, increasing the likelihood that volcanism will sample wet mantle domains. Thus, the composition of the present-day volcanic gases on Venus, that is, the proportion of water in the volcanic gas, irrespective of the total rate or volume of volcanism, links back to the water content of the interior and, critically, to the planet's climatic past.

Sources and sinks of the Venusian atmosphere

Photochemical losses

We calculated atmospheric production and destruction rates of gases using a chemical-kinetic network that reproduces the observed modern atmosphere of Venus³². Figure 2 shows the species with the highest rates of destruction in the atmosphere and their effective chemical timescales. For the atmosphere to be in a steady state, the total flux of a species X in the atmosphere, $\Delta\Phi_X$ ($\text{cm}^{-2} \text{s}^{-1}$), should be zero (Methods). The gaseous species OCS, CO_2 , and H_2O exhibit the highest rates of net destruction in the atmosphere, which is accompanied by correspondingly high production rates of other C-O-H-S species, as dictated by the conservation of elemental abundance (Supplementary Text 1 and Supplementary Fig. 1).

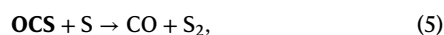
The primary reaction pathways dominating over the atmospheric column are as follows:



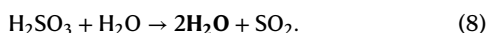
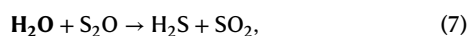
These photochemical dissociation reactions dominate at the top of the atmosphere (>90 km from the surface). In bold are the key species of each reaction, where if they are on the left hand side CO_2 , OCS and H_2O are in destruction, and if they are on the right hand side, CO_2 , OCS and H_2O are in production.

Within the clouds, the formation of sulfuric acid from SO_3 and H_2O and its degradation provide the 'seeds' for cloud formation. However, these reactions are in near-perfect balance and, therefore, contribute negligible net production or destruction of either H_2O or SO_3 .

Closer to the surface (<45 km), the following reactions dominate:



Moreover, the water chemistry is set by:



The net effect of the above reactions results in the set of dominant atmospheric destruction fluxes shown in Fig. 2. Supplementary Text 2 and Supplementary Figs. 2–4 expand further on the reaction rates.

Measurements of the atmospheric abundance of species over the 20th and 21st centuries place lower limits on the timescales for the

evolution of the planet's atmosphere (grey region in Fig. 2). Although observations of Venus's atmospheric chemical abundances have been made at different altitudes and lack a consistent methodology, they align with our self-consistent atmospheric calculations for Venus³² and are ultimately consistent (within the errors) with Venus having a broadly constant atmospheric composition over this period³³.

The key abundance ratios of species being atmospherically destroyed are OCS/ H_2O and CO_2 / H_2O . These ratios are both >1 . If these gases are being directly resupplied in these abundances, then that would indicate that the source of gas to the atmosphere is very dry. Hence, Venus would have a very dry magma if volcanism is providing the restorative flux.

However, before making the connection to volcanism and the interior of Venus, it is important to note that balancing photochemical destruction and production can resolve only the net effect of loss and resupply to Venus's atmosphere. Species undergoing net atmospheric destruction may also be part of geological cycles that contribute both further sinks (for example, by surface interaction) and sources (for example, by volcanism, as discussed later). Consequently, the photochemically derived destruction rates provide a lower bound on the resupplying gas rates into the atmosphere needed to sustain a steady state. Any other geological sinks would also need to be balanced by resupply. Most importantly for whether Venus's interior is wet, we need to consider whether there are other geological sinks of H that would mean that the required H flux back to the atmosphere is greater than we have calculated (and, therefore, that the OCS/ H_2O and CO_2 / H_2O ratios of any restorative gas flux are lower than photochemical sinks alone imply). We consider how any such sinks may affect the photochemically derived fluxes in section 'Sinks of the Venusian atmosphere'.

Sinks of the Venusian atmosphere

Data from Venus's surface and subcloud atmosphere indicate that there is continuing chemical alteration of rocks and minerals by atmospheric gases, which is another sink for some atmospheric species³⁴. Our understanding of the chemical weathering of Venus's surface is derived from mineral compositions obtained through fly-bys, orbiters and landers (for example, refs. 35–37) and the predicted stability of minerals on the surface of Venus (for example, refs. 38–40). The kinetics of these reactions are key for understanding their ability to shape the Venusian atmosphere. However, our knowledge of reaction rates is limited by the uncertain composition of the deep atmosphere and surface mineral compositions, so that there is a need for experiments under Venusian conditions. Nonetheless, we estimated the effect that surface processes have on atmospheric sources and sinks of OCS, CO_2 and H_2O , by reviewing the viability of candidate weathering reactions for each species in turn.

CO_2 . Gas species in the near-surface atmosphere could oxidize exposed igneous rocks. Surface mineral oxidation could occur through reactions with atmospheric species CO_2 , H_2O or SO_2 (ref. 41). However, given the surface temperature, pressure and composition of the Venusian lower atmosphere⁴², the equilibria of each of the redox reactions $\text{CO}_2 \rightleftharpoons \text{CO} + \frac{1}{2}\text{O}_2$, $\text{SO}_2 \rightleftharpoons \frac{1}{2}\text{S}_2 + \text{O}_2$ and $\text{H}_2\text{O} \rightleftharpoons \text{H}_2 + \frac{1}{2}\text{O}_2$ (ref. 43) are favoured to run in the direction of reducing the surface. The surface is, thus, unlikely to act as a sink for these oxidizing agents. Were oxidation to occur, then it would be a competitive reaction between these three species. Of these, CO_2 is the prime candidate as it accounts for ~96% of the surface gas-phase composition, in contrast to SO_2 at 150 ppm and H_2O at 30 ppm (refs. 37,44). CO_2 removal could, thus, occur by oxidation of pyroxene ((Ca,Mg,Fe) SiO_3) into magnetite (Fe_3O_4) and magnetite into hematite (Fe_2O_3)⁴¹. It has also been proposed that surface silicates on Venus could consume atmospheric CO_2 through carbonitization into carbonates. However, it is unclear whether this process is operating given the stability of carbonates at Venus's surface temperature⁴⁵. If carbonate–silicate equilibria are stable under Venus's

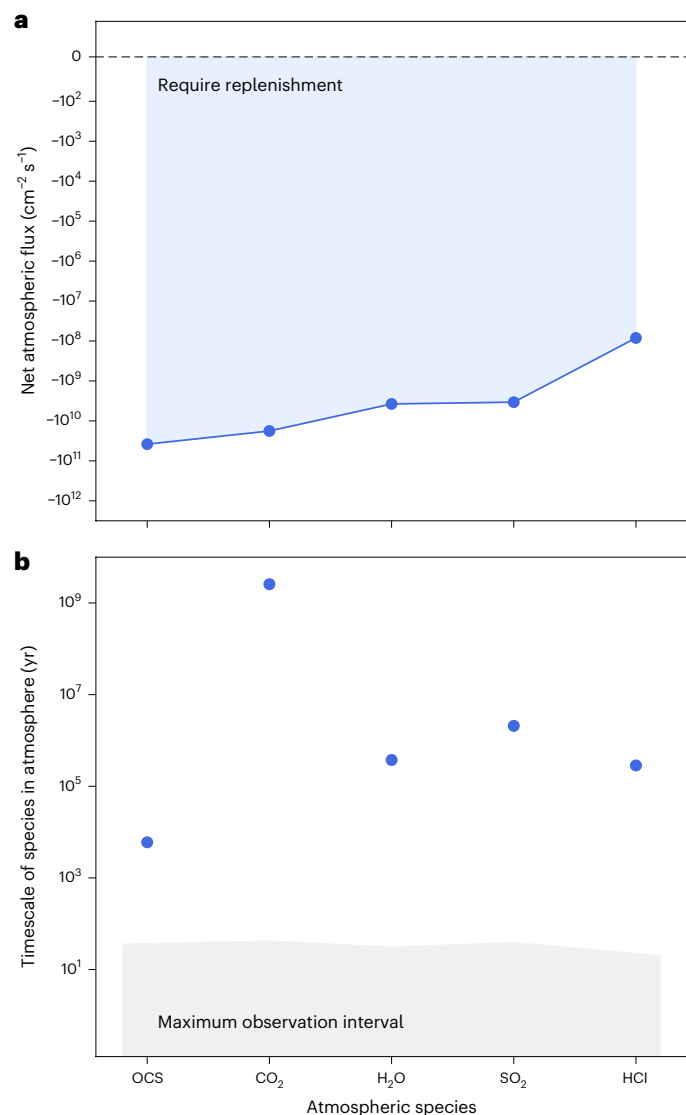


Fig. 2 | Dominant destructive fluxes of species in Venus's atmosphere. **a**, Calculated net atmospheric flux for the five species with the highest rates of destruction (blue line). Species undergoing atmospheric destruction necessitate an equivalent flux for replenishment to maintain an atmospheric steady state (blue shaded region). **b**, Timescales for species in the atmosphere, which are determined by how long it would take for the abundance of the species undergoing destruction to decrease by a factor of e (blue circles). The grey shaded area indicates the baselines of the lower limits for the stability of species in the atmosphere, which were determined by the largest time interval across observational measurements. Data sources^{37,75–78}.

conditions, they could serve as another CO_2 removal mechanism. Importantly, this analysis implies that if surface oxidation or carbonization is occurring, it would most probably require an upwards revision of the $\text{CO}_2/\text{H}_2\text{O}$ in the restorative gas, that is, towards a drier gas.

OCS. Carbonyl sulfide could also be subject to weathering sinks. It could react with the FeO in silicates, hematite, siderite (FeCO_3), magnetite or pyrrhotite ($\text{Fe}_{0.875}\text{S}$) to produce FeS_2 (ref. 41). Though the reaction kinetics are uncertain, if operating, these OCS destruction fluxes would increase the $\text{OCS}/\text{H}_2\text{O}$ ratio of the overall gas removed from the atmosphere.

H_2O . Water would be lost from the atmosphere if it were involved in forming hydrous minerals. However, no hydrous or H-bearing minerals

are stable under Venus's surface conditions (Supplementary Fig. 5). Consequently, there are no mechanisms for capturing and retaining the H from H_2O in a mineral form on the surface. Oxidation reactions with H_2O produce H_2 as a by-product, such as $3\text{FeO}(\text{s}) + \text{H}_2\text{O}(\text{g}) \rightarrow \text{Fe}_3\text{O}_4(\text{s}) + \text{H}_2(\text{g})$ (ref. 41). The H_2 can consequently serve as a reducing agent, as in $\text{H}_2(\text{g}) + \text{CO}_2(\text{g}) \rightarrow \text{CO}(\text{g}) + \text{H}_2\text{O}(\text{g})$, effectively replenishing the reacted H_2O in the atmosphere. Yet, the present-day Venusian atmosphere indicates that there is a surplus of H_2 production (Supplementary Fig. 1), necessitating a means of H_2 removal not further production. This suggests that crustal oxidation reactions are unlikely to create smaller (OCS, CO_2)/ H_2O ratios than estimated photochemically.

A second factor specific to the H budget of the atmosphere is escape to space. For a stable atmosphere, the flux of H escape needs to be balanced by a higher H content in the restorative gas flux. This would require a higher hydrogen content in the mantle than we have inferred from photochemical destruction alone. The average rate of H ion loss from Venus is $0.2\text{--}3.8 \times 10^{25} \text{ s}^{-1}$ (refs. 46–51). This rate is not enough to overcome H_2 production in the atmosphere, so that it still does not require an H_2 source due to surface reactions of H_2O . The extreme scenario for the H-loss rate in which all the escaped H is replenished by H from H_2O would require $\sim 0.2\%$ more H_2O in the replenishing gas. This remains inconsequential on the (OCS, CO_2)/ H_2O ratios, reinforcing the implication of a dry replenishing gas and, thus, a dry Venusian interior.

Sources to the Venusian atmosphere

To replenish the gases undergoing atmospheric destruction, we explored a range of possible resupply mechanisms. Only by including volcanism and ruling out other mechanisms were we left with a constraint on the Venusian interior composition.

One possibility is that exogenic sources of material resupply OCS, CO_2 and H_2O to the atmosphere. We tested this by assuming a highly favourable case for this scenario: exogeneous delivery equivalent to the terrestrial flux of $\sim 7 \times 10^{10} \text{ g yr}^{-1}$ of pure carbonaceous chondrite material⁵². This would deliver $\sim 1 \times 10^8 \text{ mol yr}^{-1}$ of sulfur, $\sim 1 \times 10^8 \text{ mol yr}^{-1}$ of carbon and $\sim 1 \times 10^9 \text{ mol yr}^{-1}$ of hydrogen⁵³. Again making the very favourable assumption that every mole of S is supplied as OCS, every mole of C as a mix of CO_2 and OCS, and every mole of H as H_2O , we calculated the maximum impact of this delivery for the composition of any endogenous atmospheric resupply. For S, the exogenic flux is four orders of magnitude smaller than the OCS destruction rate, for C, four orders of magnitude smaller than the OCS and CO_2 destruction rates, and for H, three orders of magnitude smaller than the H_2O rate. Exogenous delivery is, therefore, clearly inadequate for counterbalancing the destruction of sulfur, carbon and hydrogen in the atmosphere.

On Earth, metamorphic degassing of buried sediments resupplies (predominantly) C, S and H to the atmosphere. The importance of this process on Venus is doubtful. For metamorphic devolatilization to occur, two crucial conditions must be met: (1) mineral carriers of C, S or H must be stable at surface conditions if they are to enter the crust and (2) they must subsequently become unstable as they are buried in higher pressure and temperature environments⁵⁴. The devolatilization of H is implausible due to the lack of stable H-bearing minerals at the surface (Supplementary Fig. 5). Similarly, C-bearing minerals, like calcium carbonates, are unlikely to be stable under Venusian surface conditions⁴¹, so crustal devolatilization is unlikely to be a source of gaseous H or C species.

The story for S is more nuanced. Magnetite–pyrite ($\text{Fe}_3\text{O}_4\text{--FeS}_2$) and hematite (Fe_2O_3)–magnetite assemblages can be stable in the conditions of Venus's plains, with pyrite also potentially stable in the highlands⁴¹. It is not clear, however, the extent to which sulfidation of the crust is active. This, in part, reflects uncertainties on the thermodynamics and surface composition. However, even if S and O are present in minerals that are stable at surface conditions and their breakdown releases OCS, the process of burial necessitates active subduction, which is notably absent on modern Venus⁵⁵.

Alternative tectonic regimes for Venus have been proposed, such as the ‘plutonic-squishy lid’⁵⁶ model driven by intrusive magmatism. Although this regime might involve short-lived, sporadic episodes of ductile plate sinking, it is thought to lack the large-scale downwelling characteristic of plate tectonics⁵⁶. In the absence of planetary-scale tectonic burial, the only remaining mechanism would be volcanic burial. However, in that case, the volcanic gas released from the magmas is probably as important, if not more so, than the resultant metamorphic degassing. We, therefore, conclude that if the atmosphere is in a steady state, the gas needed to balance photochemical destruction must come from volcanism.

Volcanic gas chemistry on Venus

There is good evidence that Venus continues to be volcanically active. Venus’s surface is widely recognized as having been shaped by volcanism^{57–59}. Evidence for continuing volcanism comes from a number of sources: an observed lava flow at a volcanic vent⁶⁰ and young lava flows based on surface ageing from emissivity measurements³⁴. Moreover, volcanic degassing is necessary for maintaining the global sulfuric acid (H_2SO_4) clouds⁶¹. The key question we address here is what the average composition of this volcanism is and, in particular, its water content. The key ratios of species being photochemically destroyed, $\text{OCS}/\text{H}_2\text{O}$ and $\text{CO}_2/\text{H}_2\text{O}$, imply that the volcanic gas is very dry, and by extension, that the magma is dry. Interpreting the magma composition as a reliable proxy for the globally averaged supplying mantle, this suggests a correspondingly dry interior. We tested this hypothesis by modelling volcanic degassing speciation of a dry terrestrial basalt under Venus’s surface conditions (Methods), which allowed us to determine whether such a dry gas is consistent with Earth-like magmas.

Figure 3 (blue bars) shows the speciation of the volcanic gas degassed by dry terrestrial basalt at various Venusian pressures. In comparison to the Earth-like volcanic gas, which is characterized by a 96% molar water content degassed at Venus’s surface pressure (or a minimum of 25% when degassed at a higher pressure of 3,000 bar), we found a distinct contrast when assessing Venus’s volcanic gases from the atmospheric composition constraints, thus revealing a maximum water molar content of 6%. The $\text{CO}_2/\text{H}_2\text{O}$ abundance offset between Venus’s volcanic gas and the volcanic gas in Earth’s mid-ocean ridge basalt (MORB) decreased with increasing degassing pressure (Fig. 4), although it is critical to note that the Venusian values represent lower limits, as we have not accounted for potential weathering sinks of OCS and CO_2 (see section ‘Sinks of the Venusian atmosphere’). The $\text{OCS}/\text{H}_2\text{O}$ shows the largest discrepancy with terrestrial basalt, being at least four orders of magnitude too low in the terrestrial magma, indicating four orders of magnitude more H_2O . This considerable offset in abundances in the volcanic gases indicates that the Venusian magma lacks H_2O , instead being sulfur and carbon rich.

If the restorative gas is supplied through volcanism, it is probably degassed at high temperatures and pressures and is, thus, in local thermodynamic equilibrium. By taking a system with compositions defined by OCS, CO_2 and H_2O in their inferred relative proportions, we varied the pressure and temperature to examine whether these species can be in thermochemical equilibrium in their inferred abundances (Methods). We found that the inferred $(\text{OCS}, \text{CO}_2)/\text{H}_2\text{O}$ ratios matched, within an order of magnitude, for a temperature of $\sim 1,890$ K and a pressure of ~ 60 bar (Fig. 3). This alignment suggests that the observed flux ratios originated from a high-temperature process, consistent with volcanism.

Implications for the climate histories of Venus-like planets

A dry Venusian interior is not consistent with Venus having had surface oceans or, by extension, a conventionally habitable climate. The most plausible way for both the atmosphere and interior to have efficiently lost the originally accreted H is if the water reservoir existed as a steam

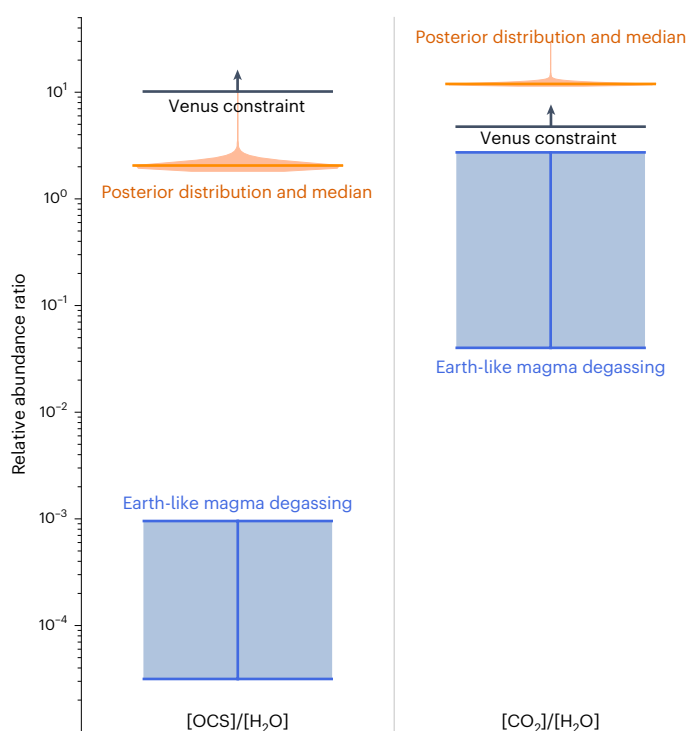


Fig. 3 | Constrained Venusian volcanic gas compared with Earth-like magma degassing. OCS and CO_2 abundances relative to H_2O abundance ($\text{OCS}/\text{H}_2\text{O}$ and $\text{CO}_2/\text{H}_2\text{O}$, respectively) in Venus’s restorative volcanic gas. These ratios are constrained directly from atmospheric destruction fluxes of OCS, H_2O and H_2O (black lines). The relative abundances provide a conservative lower limit for the volcanic gas speciation due to the potential surface weathering sinks affecting OCS and CO_2 , necessitating a higher restorative degassing flux to balance them, although H_2O is unlikely to be affected by surface sinks from weathering. The posterior distributions were obtained from a Markov chain Monte Carlo sampler by considering variations in degassing pressure and temperature (represented by orange violins). The horizontal orange lines denote the medians (‘Bayesian sampling for volcanic gas speciation at thermochemical equilibrium’). This median signifies the equilibrium gas composition that best aligns with the atmospheric speciation constraints. The blue bars indicate the range of $\text{OCS}/\text{H}_2\text{O}$ and $\text{CO}_2/\text{H}_2\text{O}$, when degassing a terrestrial basalt under different Venusian pressure conditions, ranging from the deep interior to the surface pressure (‘Modelling the volcanic degassing of a dry terrestrial basalt’).

atmosphere above a magma ocean for tens of millions of years, thus leaving a long-lived dry surface¹³.

D/H ratios have been a historically important constraint on the water history of Venus⁵. The climate history we infer here puts this central observation in a new light. The long-term desiccation of the planet is most consistent with a steady-state source of continuous outgassing to the atmosphere⁶². The D/H ratio of the water being outgassed would have to be substantially higher than that of terrestrial water. This is consistent with degassing from a highly fractionated mantle source, as would be the case if the mantle inherited the fractionation from the early coupled atmosphere–magma ocean loss episode¹³, when the planet first dried out.

These results indicate that Venus probably never experienced conditions conducive to ocean condensation. This result is in agreement with recent modelling work indicating that a Venus born with a hot steam atmosphere would never have received a low enough insolation to condense its water reservoir⁶³. Consequently, Venus-like exoplanets or planets within the Venus zone⁶⁴, which the James Webb Space Telescope can characterize, are unlikely to be cool enough to condense liquid water if they formed in situ. This makes these planets improbable candidates for liquid-water habitable conditions.

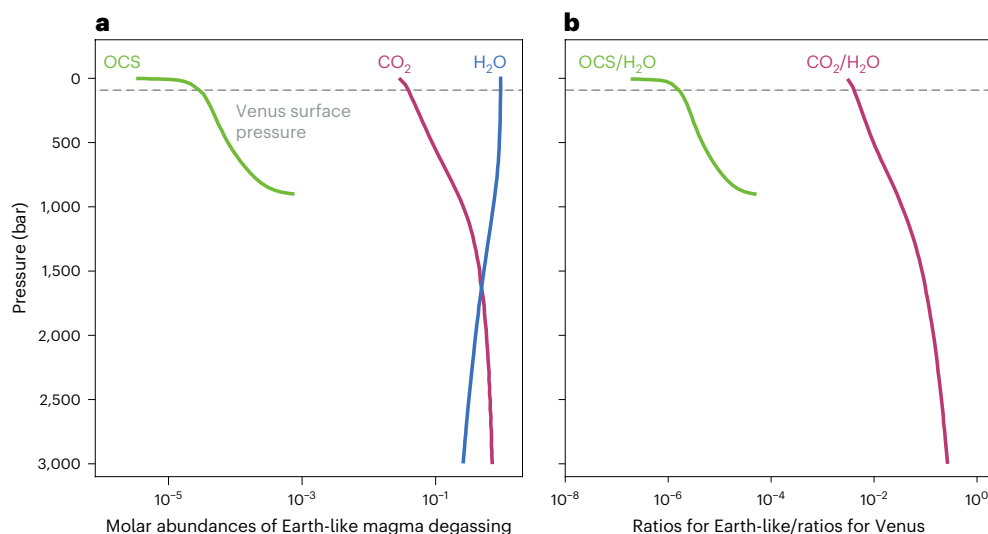


Fig. 4 | Comparison of volcanic gas compositions across pressure. **a**, Variation in MORB degassing speciation under different pressure conditions, plotted against the volcanic gas molar abundances of OCS and CO₂ relative to H₂O. **b**, Ratios of the relative molar abundances OCS/H₂O and CO₂/H₂O degassed from

Earth's MORB to OCS/H₂O and CO₂/H₂O from Venus's volcanic gas, as discussed in 'Volcanic gas chemistry on Venus'. Venus's surface pressure is plotted for reference (black dashed lines).

Methods

Modelling the Venusian atmosphere

In modelling the atmosphere, a one-dimensional Lagrangian high-energy photochemistry-diffusion model⁶⁵ called ARGO was used as a numerical solver for STAND2022, an atmospheric chemical network describing gas-phase photochemical kinetics^{32,65}. STAND includes over 6,600 thermochemical and photochemical reactions involving 480 species composed of C-O-H-S-N-Cl and a handful of other elements. The chemical kinetics code has been validated to within an order of magnitude of most observations of atmospheric species abundances for Venus³².

The reactions from STAND were solved by ARGO as a set of time-dependent, coupled, nonlinear differential equations:

$$\frac{\partial n_X}{\partial t} = P_X - L_X n_X - \frac{\partial \Phi_{\text{diff},X}}{\partial z} \quad (9)$$

at height z (cm) from the surface, where n_X (cm⁻³) is the number density of a species X , P_X (cm⁻³ s⁻¹) is the rate of production of species X , L_X (s⁻¹) is the rate constant for destruction of species X and $\Phi_{\text{diff},X}$ (cm⁻² s⁻¹) is the vertical flux of species X as a function of eddy- and molecular-diffusion processes. P_X and L_X were determined from the rate constants of the different reactions within STAND and from the relevant species abundances. $\partial n_X / \partial t$ represents the rate of change of number density of species X at each height z in the atmosphere and is a function of the species production and destruction and the change of the species vertical flux with respect to height.

These chemical reactions solved by ARGO determine the chemical production and loss rates of each species at each altitude level in the atmosphere. ARGO traced the path of a gas parcel as it moved from the surface to the top of the atmosphere and back down. It started with initial chemical composition data at the base of the atmosphere. At each altitude during the upward journey, ARGO solved equation (9) for all atmospheric species using pressure, temperature and species abundances. The time interval for solving was determined by the eddy-diffusion profile, which accounted for vertical transport. The pressure-temperature and eddy-diffusion profiles for modern Venus were adopted from previous photochemical kinetics models of the lower and middle atmosphere (for more details, see ref. 32 and references therein). At the top of the atmosphere, ARGO incorporated

the incident stellar spectrum and included photochemical reactions driven by the stellar flux during the downward journey. The process was iterated until a convergence criterion was met, leading to a comprehensive global solution.

Atmospheric flux and timescale measurements

The net photochemical and thermochemical rate of production or destruction of X in the atmosphere, $\Phi_{\text{atmo},X}$ (cm⁻² s⁻¹), was used to determine the chemistry of the restorative gas flux in the atmosphere ('Photochemical losses'). $\Phi_{\text{atmo},X}$ was calculated for each species X by summing over the atmospheric chemical production and destruction of X at each altitude step at height z , as determined by ARGO, where

$$\Phi_{\text{atmo},X}(z) = \Phi_{\text{photochem},X}(z) + \Phi_{\text{thermochem},X}(z) + \Phi_{\text{diff},X}(z), \quad (10)$$

where $\Phi_{\text{photochem},X}$ (cm⁻² s⁻¹) is the photochemical rate for X and $\Phi_{\text{thermochem},X}$ (cm⁻² s⁻¹) is the thermochemical rate for X . The change in number density is

$$\frac{dn_X(0,t)}{dt} = \frac{d\Phi_X(t)}{H_{\text{sc}}} \quad (11)$$

at surface pressure, where Φ_X (cm² s⁻¹) is the *total* species flux in the atmosphere and H_{sc} (cm) is the atmospheric scale height. Assuming a steady state, each species has a fixed number density, $dn_X(0,t)/dt = 0$, so that $\Delta\Phi_X$ is zero for all X . Within our atmospheric model, the bottom of the atmosphere was incorporated as fixed, initial, surface-mixing ratios. These were chosen so that the model fitted observational measurements of species abundances further up in Venus's atmosphere³².

Atmospheric processes alone are not sufficient to describe all processes affecting atmospheric gas fluxes Φ_X , as the model does not include interactions with the surface or atmospheric escape. The net atmospheric flux is

$$\Delta\Phi_X = \Phi_{\text{deg},X} - \Phi_{\text{esc},X} - \Phi_{\text{sink},X} + \Phi_{\text{atmo},X} = 0, \quad (12)$$

where $\Phi_{\text{deg},X}(t)$ (cm⁻² s⁻¹) is the surface volcanic degassing of X , $\Phi_{\text{esc},X}$ (cm⁻² s⁻¹) is the escape flux, $\Phi_{\text{sink},X}$ (cm⁻² s⁻¹) is the deposition flux due to interactions with the surface and $\Phi_{\text{atmo},X}$ (cm⁻² s⁻¹) is the net photochemical and thermochemical production or destruction of X in the atmosphere, as calculated by our model.

This was used to constrain $\Phi_{\text{deg},X}$ to gain insights into the volcanic degassing speciation on Venus. There may still be a redundancy, however, in solving for equation (12), as the weathering reaction rates $\Phi_{\text{sink},X}$ for the Venusian surface conditions are still mainly unknown, as addressed in ‘Sinks of the Venusian atmosphere’.

The $\Phi_{\text{atmo},X}$ fluxes (Fig. 2, top) integrate photochemical, thermochemical and diffusive processes within the atmosphere:

$$\Phi_{\text{atmo},X} = \int_0^\infty \left[P_X(z, t) - L_X(z, t)n_X(z, t) - \frac{\partial \Phi_{\text{diff},X}}{\partial z} \right] dz. \quad (13)$$

To extract $\Phi_{\text{atmo},X}$ from the model, we tracked the history of the air parcel throughout the cycle by calculating the change in molecular abundance in the parcel at the start of the gas parcel cycle compared to the end (Δn) when convergence had been reached and the parcel had reached the surface again:

$$\Phi_{\text{atmo},X} = \frac{2\Delta n K_{ZZ}}{H_{\text{sc}}}. \quad (14)$$

The coefficients used to construct P_X and L_X were provided by the chemical network STAND. The velocity of the gas parcel in the model was determined through $2K_{ZZ}/H_{\text{sc}}$, where K_{ZZ} is the eddy diffusion at the surface ($\text{cm}^2 \text{s}^{-1}$).

The e-folding time of atmospheric species was determined from their expected lifetime in the atmosphere (Fig. 2, bottom), defined as the species column density divided by $\Phi_{\text{atmo},X}$.

The most dominant reactions for each species were determined by comparing the column rates R_{chem} ($\text{cm}^{-3} \text{s}^{-1}$) of individual reactions (as plotted in Supplementary Figs. 3–5). At the last time instance that a parcel spends at each height z ,

$$R_{\text{chem}} = k(z) \prod_{i=1}^N n_{X_i}, \quad (15)$$

where $k(z)$ ($\text{cm}^{3N-3} \text{s}^{-1}$) is the reaction rate constant extracted from STAND (right before the parcel moved; Supplementary Text 2) and N is the number of reactants.

Modelling the volcanic degassing of a dry terrestrial basalt

The degassing speciation of a dry terrestrial basalt was simulated using a thermodynamic magma degassing model, EVO^{66–68}. A MORB served as a valuable reference point for the ambient mantle of a plate-tectonically active planet, far away from direct inputs of wet material at subduction zones. EVO iteratively calculated the melting and subsequent volatile outgassing of a gas in equilibrium with its parent, a MORB magma. The melting model within EVO used the batch melting equation⁶⁶ to partition volatiles from the bulk mantle into the melt phase during each time step.

This model calculated the speciation and volume of a C–O–H–S gas phase, including our key species OCS, CO_2 and H_2O , in equilibrium with a silicate melt at a given pressure, temperature and magma oxygen fugacity (f_{O_2}), considering both homogeneous gas-phase equilibria and heterogeneous gas-melt equilibria considered in the form of solubility laws suitable for a melt with a basaltic composition (detailed across refs. 66–68). The only exception was for OCS, as its abundance was calculated in another gas-phase thermodynamic equilibrium step, because there was a lack of solubility data for magmas for it to be considered in a melt-gas equilibrium.

EVO was initialized with volatile contents characteristic of a MORB: 0.3 wt% H_2O and 0.01 wt% total sulfur (as a sum of dissolved SO_2 and H_2S). The f_{O_2} was set to a fayalite-magnetite-quartz buffer (FMQ) of 0.5, which aligns well with the range observed in erupting MORBs on Earth (FMQ = 0.41 ± 0.43)⁶⁹. The model simulated degassing at a fixed temperature of 1,400 K (representing the average temperature of degassing

magma) and a pressure range spanning beyond potentially relevant Venusian conditions, from deep within the interior at 3,000 bar to the surface pressure on Earth 1 bar.

Surface-atmosphere thermochemical equilibrium

We examined the stability of H-bearing minerals at the surface using GGCHEM⁷⁰ by considering a mixture of gas and condensed species at 92 bar with surface elemental abundances estimated from surface oxide ratios measured by Vega 2 and the surface gas abundances in our atmospheric model (as in ref. 32). Supplementary Fig. 5 provides insight into the stable H-bearing condensates across a temperature range, encompassing both the cooling of molten magma to Venus’s surface temperature and extending to lower temperatures where solid phases become more stable. At Venus’s surface conditions, H exists solely in gaseous form (blue line in Supplementary Fig. 5). Water in the magma is unstable under Venus’s surface conditions and preferentially exsolves (Fig. 4). Even if water were to remain in the magma after degassing, it would eventually diffuse out, potentially becoming a long-term source of water in the atmosphere, as the absence of stable hydrous or H-bearing minerals indicates that the surface cannot retain hydrogen. This gaseous release into the atmosphere would still be included in the resupplying H_2O flux we constrained atmospherically. Therefore, based on the understanding that we could not reproduce the atmospherically constrained volcanic water content through magma degassing and that water cannot be retained in the surface within solids or minerals, we conclude that the surfacing magma itself must be dry.

Bayesian sampling for volcanic gas speciation at thermochemical equilibrium

We investigated the closest-fitting composition of the atmospherically inferred volcanic gas to a gas in thermochemical equilibrium using FastChem⁷¹. Using the elemental abundances inferred from the proportions of OCS, CO_2 and H_2O , we varied the pressure and temperature to determine whether these species can be found in thermochemical equilibrium in their inferred abundances. We used the emcee affine-invariant ensemble sampler for Markov chain Monte Carlo sampling⁷² to sample the pressure and temperature priors for the volcanic gas and determine the best fit for our atmospheric constraints. χ^2 was used as a measure of the goodness of fit between the gas speciation inferred from our atmospheric model to the equilibrium gas speciation produced by FastChem.

The priors were defined by the range of pressures over which a magma can be degassed, from deeper within the planetary interior at 3,000 bar to pressures higher up in the atmosphere at 60 bar. Similarly for the temperature, which ranged from 1,900 K to a cooler 1,000 K as the magma cooled down. Although the most probable priors were at the edge of the prior distribution, suggesting that further extremes might be a better fit, extending the distribution would not correspond to feasible degassing conditions. The complete resulting gas chemistry of the most probable gas was primarily composed of CO , CO_2 , S_2 , OCS, H_2S and H_2O . This suggests that the constraining speciation ratios used in the χ^2 measure of fit would differ following the incorporation of weathering sinks of the key species (as shown by the upward arrows in Fig. 3, which indicate that weathering would shift the fluxes to drier ratios).

Data availability

No new observational data were generated for this work. The data used to produce Figs. 2 and 3 related to Venus’s atmospheric chemistry can be found across refs. 32, 65, 73, 74. The data needed to evaluate the findings have been deposited in the Harvard online database at <https://doi.org/10.7910/DVN/19HF6Q>.

Code availability

The methods underlying ARGO and STAND are comprehensively documented in a collection of publicly available papers^{32, 65, 73, 74}. The ARGO

software with the atmospheric flux calculation extension may be made available upon request. The EVo version used here is publicly available via GitHub at <https://github.com/pipliggins/EVo> and is documented across refs. 66,67. The FastChem 2.0 model⁷¹ is freely available via GitHub at <https://github.com/exoclime/FastChem>, where updates are hosted. GGChem⁷⁰ is also freely available via GitHub and updated at <https://github.com/pw31/GGchem>.

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Author contributions

T.C., O.S. and P.B.R. conceptualized the study. T.C. performed the modelling and analysis. O.S. and P.B.R. supervised the project. All authors contributed to the writing of this paper. P.B.R. focused their contribution on the atmospheric chemistry aspects of the paper.

Competing interests

The authors declare no competing interests.

Additional information

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